COLD DOWNFLOWS OF GROUNDWATER AT OHAAKI GEOTHERMAL FIELD: PRELIMINARY RESULTS

C.J. BROMLEY[1], T.M. HUNT[1] AND C. MORRIS[2]

[1]Institute of Geological and Nuclear Sciences Ltd., Wairakei
[2]Geothermal Group, Electricity Corporation of New Zealand, Wairakei

SUMMARY - An exploitation-induced downflow of cold groundwater in the northwestern part of the field is inferred from localised changes in groundwater level. Geophysical techniques (GPR, VES, SP) were tested to evaluate their ability to map hot and cold groundwater surfaces, their changes with time, and to determine the direction of fluid flow in the groundwater zone. None of the techniques were found to be entirely adequate for these purposes. Data from 17 new, shallow monitor holes suggests that shallow, perched groundwater aquifers, of limited horizontal extent, overlie the regional groundwater near several areas of thermal activity, and that production from the field has resulted in them being partially drained. The data also suggest that the regional groundwater level has been depressed by up to 5 m within about 100 m of **BR4.** The original concept of a single, extensive cone of groundwater depression centred near **BR4,** should be replaced by the concept of several, small, localised depressions.

1. INTRODUCTION

Many liquid-dominated geothermal fields under exploitation have cold downflows. Evidence for cold downflows generally comes from measurements in shallow production wells which show declines (with time) in feed temperatures, chloride concentration, and enthalpy. These data, together with falling groundwater levels in the vicinity, suggest that significant amounts of cold (< 100°C) water are draining down from the overlying groundwater zone into the upper part of the geothermal reservoir. Internal downflows can also occur in drillholes associated with damaged well casing, or behind imperfectly cemented casing.

Downflows have been reported at Wairakei (Grant, 1981; Allis 1980; Bixley, 1990), Kawerau (Grant et al, 1982; Bixley, 1990), Tiwi (Alcaraz et al, 1989), and Bulalo Geothermal Fields (Villadolid, 1991). The effect of downflows can range from being little more than a nuisance to a major problem which seriously reduces production and causes expensive production wells to be "written off" (such as at Tiwi and Wairakei).

At Wairakei, a cold downflow has affected the northeastern part of the Western Borefield. Evidence includes a decline of about 100"C in feed temperatures at shallow depths (+50 to +200 masl), between 1958 and 1990. There is also evidence of chemical dilution in affected well discharges (from 1585 mg/kg chloride in 1961 to 1437 mg/kg in 1982), and of decreasing groundwater levels (by as much as 30 m; Allis, 1982). The distribution of cool fluid indicates that the downflow is a tongue, dipping gently from near Geyser Valley in a south-westerly direction (Grant, 1982; Electricorp, 1990); by the early 1980's the tongue was about 500 m wide and at least 1 km long (Grant 1982). In the late 1970's it was realised that a significant part of the downflow was associated with vertical flows in several non-producing wells which had casing breaks. These were sealed off and since then the rate of temperature decline in liquid production wells in the area has reduced (Electricorp, 1990). Allis (1982) estimated that cold water infiltration was about 200 ± 100 kg/s in 1979, including about 100 kg/s of internal cold flows in wells, which leaves approximately 100 kg/s of natural cold downflow.

Few studies published to date have specifically addressed the interaction between a geothermal system and the overlying cold groundwater system, or the mechanism of downflows. At present we know that downflows:

- occur in many liquid-dominated fields under exploitation;
- are often associated with changes in the behaviour of surface thermal features (geysers, fumaroles, hot springs, thermal ground);
- may be associated with active faults, or hydrothermal eruption breccia pipes, which provide a high permeability path between the reservoir and the groundwater zone;
- do not appear to be directly related to ground subsidence;
- are sometimes associated with a local cone of depression in the groundwater surface.

2. CONCEPTUAL MODEL

From the above data a conceptual model for the origin of downflows in liquid-dominated fields has evolved (Allis, 1982; Grant et al, 1982; Hunt and Bibby, 1992). Current models for liquid-dominated geothermal reservoirs in their natural state envisage a vertical plume of near boiling fluid rising slowly under buoyancy forces, from depths of greater than 5 km. Near the surface, the plume meets large volumes of cold groundwater, and the resultant cooling causes precipitation of silica which reduces permeability of the rocks, and restricts upward flow. In places where good vertical permeability occurs, such as active faults, the fluid rises readily towards the surface, mixes with and heats the groundwater, and emerges in a variety of thermal features.

An initial effect of exploitation of a liquid-dominated geothermal reservoir is a reduction in pressure near the top of the plume which causes the formation, or extension, of a two-phase zone. The relative mobility of steam is much higher than that of water, so there may be an initial increase in natural steam flux towards the surface, causing a heat pulse. Continued exploitation may cause vertical (downward) expansion of this zone, decreases in liquid saturation in the upper part of the zone, and further reduction in pressures. As the pressures decline, the amount of fluid passing up through the groundwater zone decreases, and there is an associated fall in the temperatures and flow rates of the discharge features. If pressures in the upper part of the two-phase zone fall sufficiently, then the upward flow of fluid may cease and the discharge features die. If pressures fall further, then the flow may reverse and cold groundwater will flow down into the geothermal reservoir. The groundwater will flow down or along the highest permeability path, condensing steam and cooling liquid in the vicinity, and becoming heated until it reaches thermal equilibrium.

For field management purposes it would be desirable to anticipate where downflows are likely to occur and to be able to minimise them if they do occur. To do this it is necessary to know more about their physical properties and to have a mathematical model for a downflow system. In particular, we need to know:

- how to locate them;
- why they are located where they are - not all upflow paths turn into downflows;
- what are the dynamic hydrological properties of the system, such as downflow rate and size of the "drainpipe".

We report here the preliminary results of a long-term investigation of cold downflows aimed at understanding and quantifying the processes involved.

We have begun to study one such downflow located at the Ohaaki Geothermal Field; this one was chosen mainly because it is comparatively well-documented, and access is relatively easy, As a first step, we have concentrated on testing geophysical methods that might help determine the location, magnitude and extent of changes in groundwater level, and thus locate the downflow.

3. DOWNFLOW AT OHAAKI

A downflow exists in the north-western part of the Ohaaki (Broadlands) Geothermal Field in the vicinity of well BR4, a deep well which is not used for production (Fig. 1). Evidence for this downflow rests largely on changes in the groundwater level, which in this area is about 300 $\pm$ 10 m (R.L. = sea level).

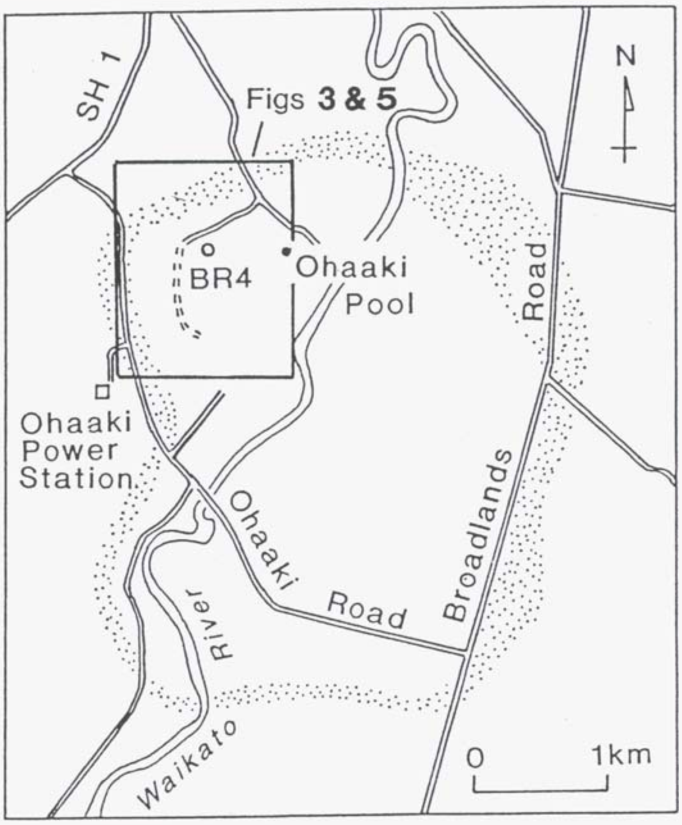

Fig. 1 Location map of Ohaaki Geothermal Field. The dotted zone indicates the resistivity boundary of the field.

During pre-production discharge tests in 1968 - 1971, water levels in shallow monitor holes 3/0, 4/0, 9/0, 11/0 and in the Ohaaki Pool fell by up to 7 m, and subsequently rose slowly after the tests finished (Hunt, 1987). These groundwater level changes followed deep-liquid pressure changes in the reservoir which fell by up to 2 MPa (1971) and subsequently rose by about 1.3 MPa (by 1975) (Grant et al, 1982).

Groundwater levels again fell in this area after production from deep wells in the field began in 1989; the largest recorded change was in BR4/0 (shallow groundwater monitor well near BR4), where the level dropped by about 10 m (from 13 m to 23 m depth) in the 3 years between early 1989 and the end of 1991 (Fig. 2). Deep water levels in the geothermal reservoir beneath the north-western production area have declined from about 250 m (R.L.) in mid 1988 to about 140 m (R.L.) at the end of 1991 and 120 m (R.L.) in early 1993 (Sherburn et al, 1993); this corresponds to deep-liquid pressure drops of about 0.9 and 1 bar respectively. The changes in groundwater level were well in excess of changes in monitor holes elsewhere in the field (<2 m), and appeared to define an elongated cone of depression in the groundwater surface suggesting the presence of a cold downflow centred near **BR4.** In some monitor holes, the water temperature (near the groundwater surface) rose soon after production began but then

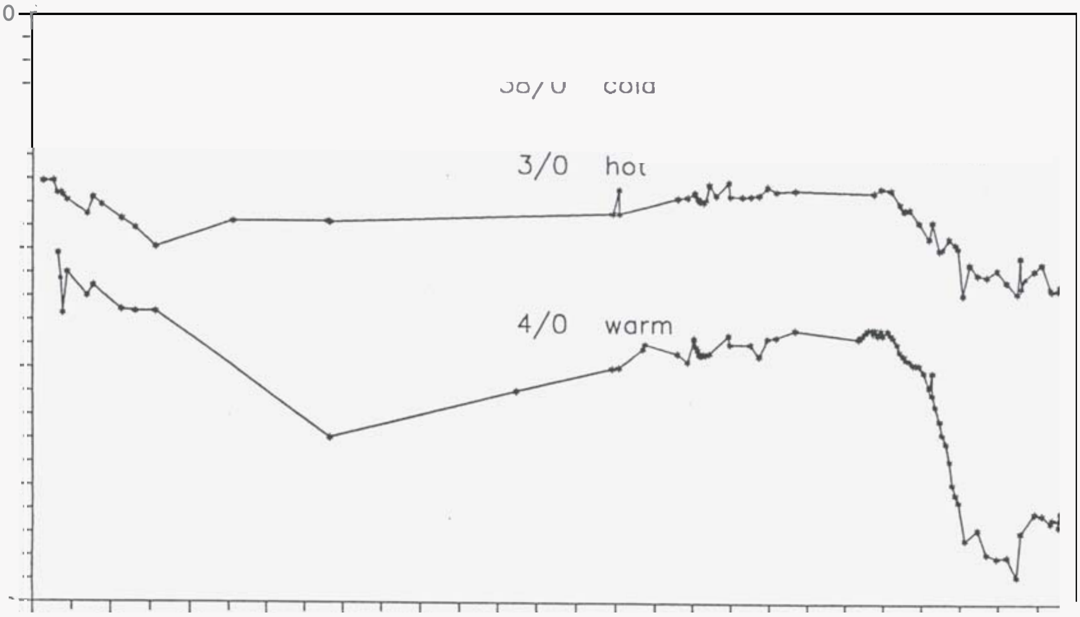

Fig. 2 Water levels of selected groundwater bores in north-west part of Ohaaki from 1967 to 1993.

decreased, suggesting a heat pulse caused by increased steam mobility followed by a decline in the upflow of steam. Temperature data in some nearby deep wells have shown evidence of cooling since production began.

It was on the basis of the above data that in mid-1992 we decided to test various geophysical methods in the vicinity of BR4.

4. GEOPHYSICAL INVESTIGATIONS

4.1 Ground Probing Radar

Initially, we thought that ground probing radar (GPR) would be the best tool to map the groundwater surface in the vicinity of BR4, and to determine the extent of the depression in the groundwater level. The advantages of GPR are rapid, inexpensive data acquisition, and high resolution of shallow electrical conductivity/permittivity interfaces. However, the method has limited penetration depth in a conductive media. Previous electrical resistivity measurements in the area had focused on providing data at several hundred metres depth; little was known about the resistivity or conductivity at shallower depths. We decided, therefore, to first conduct a number of shallow-penetration, Schlumberger-type resistivity soundings (VES) (maximum AB/2 = 200 m) in the area, before committing ourselves to the use of GPR (Fig. 3). The results from these soundings showed that, near a thermal area adjacent to BR4, resistivities of about 30 ohm-m, underlain by less than 5 ohm-m, occur within a few metres of the surface (Fig. 4). We therefore considered it unlikely that the GPR method would be able to detect the groundwater table at depths of up to 25 m near BR4.

4.2 Electrical Resistivity Soundings

Instead of using GPR, further resistivity soundings were made. The resistivity soundings were interpreted in the standard manner using one-dimensional inversion to obtain a best-fitting, layered model.

The resistivity models from 6 soundings, projected onto an E-W cross-section together With water level and temperature data from adjacent bore holes (Fig. 4), show that at most sites outside the thermal area a large decrease in resistivity occurs at about 1 to 2 metres above the groundwater level (as measured in April 1993). We interpret this decrease to be caused by the change in resistivity between material in the dry or partly saturated vadose zone, and the saturated groundwater zone. The differences of 1 - 2 m in depth between the values of groundwater level obtained from the soundings and those from measurements in monitor wells is not considered to be significant because of the assumptions and uncertainties inherent in the modelling. Furthermore, most of the soundings were made in October 1992 (soundings A - H) and water levels rise and fall by about 1 metre due to seasonal rainfall changes. *Also,* it is likely that partial saturation in the vadose zone immediately above the unconfined groundwater surface will result in an electrical conductivity contrast occurring slightly above the water table.

Soundings within the area of thermal ground east of BR4 suggest there is a layer of about 30 ohm-m resistivity, located 10 to 15 m above the main hot groundwater surface (characterised by less than 5 ohm-m). This intermediate resistivity layer we interpret to be caused by hydrothermal clay alteration and localised, perched, warm groundwater aquifers in elevated areas which are recharged by both rainwater and steam condensate. During times of high steam flux and/or rainfall recharge, these local perched aquifers may intermittently control the water levels in the monitor bores; at other times these aquifers drain into the deeper, regional, groundwater aquifer.

We infer from the data that resistivity soundings are a useful tool for approximately determining the depth of a cool groundwater surface, and of a steam-heated groundwater surface, but may not be an appropriate technique for monitoring changes in groundwater level, because of the effect of residual saturation and clays in the vadose zone.

4.3 Self-potential survey

A self-potential (SP) survey was conducted during January 1993 in the vicinity of BR4 in an attempt to locate anomalies caused by vertical or horizontal fluid flows (streaming potentials). The SP method has been used at Rotokawa and Mokai geothermal fields (Hochstein et al, 1990) in an attempt to locate hot upflows, which feature as large scale (>200 mV amplitude, >1 km wavelength), positive anomalies. Generally, however, the SP effects of thermally induced flows cannot be separated from streaming potentials caused by lateral, terrain-induced, groundwater flows.

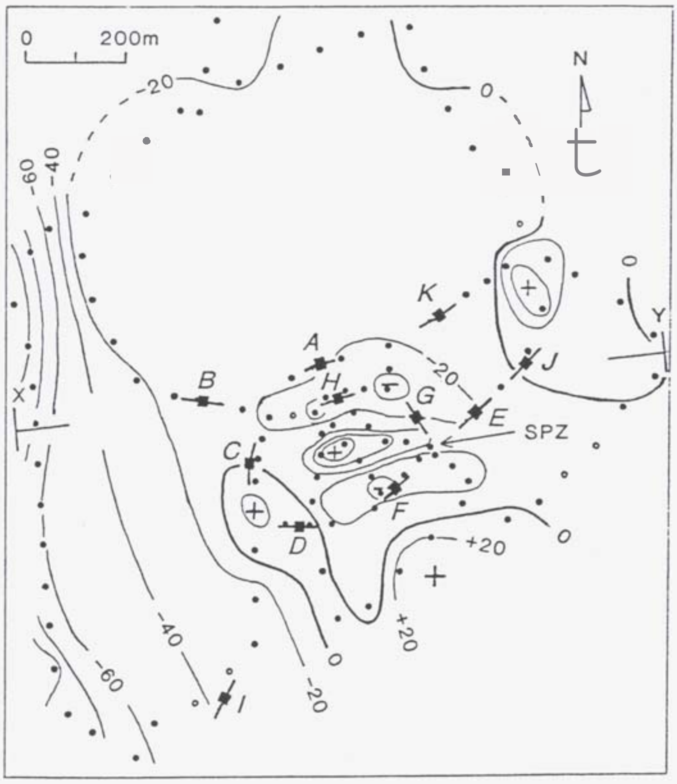

Fig. 3 Self potential anomalies in the north-western part of Ohaaki field. Contour interval is 20 mV. SP station locations are marked by solid dots; anomaly values are in terms of a zero point at SPZ. Resistivity sounding locations (**A** - J) are shown by solid squares. **X-Y** marks the position of the cross-section given in Fig. 4.

The measurements at Ohaaki were made using a simple system of 50 m or 100 m dipoles to take consecutive, differential voltage measurements around closed, inter-connected loops. Closure errors were generally less than 5 mV, however, repeat measurements on different days and with overlapping loops, showed that the repeatability is about ± 15 mV. The measurements were reduced in terms of an arbitrary base station near bore WLB 11 (SP2; Fig. 3). **In** places, strong local disturbances were measured which are probably the result of electrodes being located in hot ground, or close to pipelines; such data has been disregarded in the preparation of a smoothed contour map of the **SP** anomalies (Fig. 3).

The map (Fig. 3) shows a trend of increasing SP downslope from Ohaaki Road towards a swampy area north of BR4, consistent with terrain-induced groundwater flow. In the thermal area east of BR4, shorter wavelength anomalies predominate. SP values are negative near BR4 and WLB 12, possibly indicating a net downflow of groundwater in that area. SP values are positive near the high elevation ground close to WLB 13, possibly caused by an upflow of steam. Large, lateral, SP gradients indicate that localised, near-surface effects are probably predominant. This makes interpretation particularly difficult, especially in areas with limited coverage. However, the SP profile (Fig. **4)** illustrates that there is an approximate correlation between SP anomalies and groundwater levels, which are indicative of vertical and/or horizontal flows.

4.4 New groundwater boreholes

Seventeen shallow boreholes (WLB 1 - 17) were drilled to depths of about 20 m, in the vicinity of BR4 (Fig. *5)*. Borehole WLB 17 was cored at 11.5m - 13 m and 17 m - 18.3 m. The rocks below the water table in this bore are mainly partially-consolidated alluvial sands, interspersed with bands of clay. Cuttings were sampled over a range of depths from most boreholes and suggested similar lithologies. The bores were completed with 50 mm PVC tubing; the bottom 6 m joint being fully slotted.

A contour map (Fig. 5) of the groundwater surface, derived from the water levels in these 17 boreholes and several existing groundwater boreholes, suggests there is a west-east regional groundwater gradient and that cold groundwater is probably flowing from higher elevation ground (near 38/0) down towards the Waikato River.

Superimposed on this regional groundwater trend are pockets of elevated (by up to 6 m) water level near bores WLB 10 and WLB 11, which are located close to areas of actively steaming ground. These are interpreted to indicate the presence of perched aquifers of limited extent and thickness, which are recharged by steam condensate and rain water.

Boreholes WLB 1-17 were drilled between 25 and 31 March 1993; within a week the water levels in most boreholes had fully stabilised. However, between 8 and 22 April the level of the water in WLB 12 dropped by 10 m from 307.6 m to 297.3 m (R.L.), and temperatures increased by about 4°C. This is consistent with a perched aquifer which initially controlled the water level, draining away around the outside

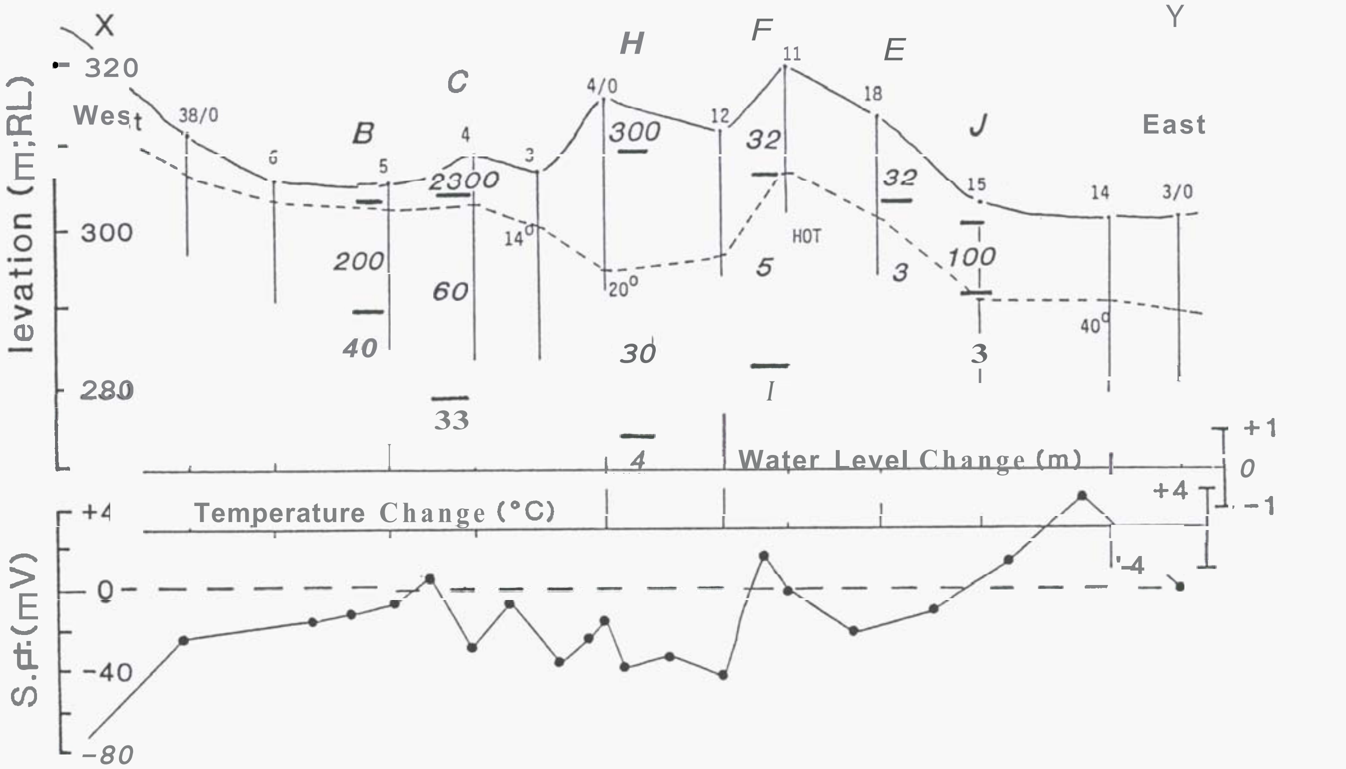

Fig. 4 West-east cross-section showing interpretations of resistivity soundings (letters), ground water levels (----) on 22/4/93, and temperatures (°C). Water level and temperature changes over a 3 month period are plotted beneath the respective boreholes. SP anomalies (mV) along the profile are also plotted; location of the profile is shown in Fig. 3.

of the casing, leaving the deeper and hotter groundwater aquifer in control of the water level.

Near boreholes BR4/0 and WLB 12 the groundwater level is depressed (by up to 5 m).

Water level and temperature changes over a 3 month period from 22 April (autumn) to 27 July 1993 (winter) showed that in most boreholes tapping the cold groundwater there was a water level rise averaging 0.5 m, and a temperature *drop* of 1°C; this is likely to have been caused by rainfall recharge. However, BR4/0 and WLB 12 showed water level rises, together with a 4°C temperature *increase*; this is interpreted to be caused by an increase in the downward percolation of stem-heated rainfall recharge to the groundwater aquifer. Most hot wells showed only minor changes, whereas warm well WLB 10 revealed a 2 m drop in its elevated (perched) water level along with a *2°C* temperature drop. Repeat measurements at most wells, taken daily in early August 1993, showed a rapid response to moderate rainfall, with water level increases of up to 9 cm.

5. DISCUSSION

In mid-1967, prior to exploitation, the groundwater level in BR4/0 was at about 304-306 m, approximately 5 m *above* the present regional groundwater level. Now (mid-1993), the groundwater level in BR4/0, and in nearby WLB 12, is depressed by up to 5 m *below* the regional groundwater level. The water levels in WLB 10 and WB11, however,

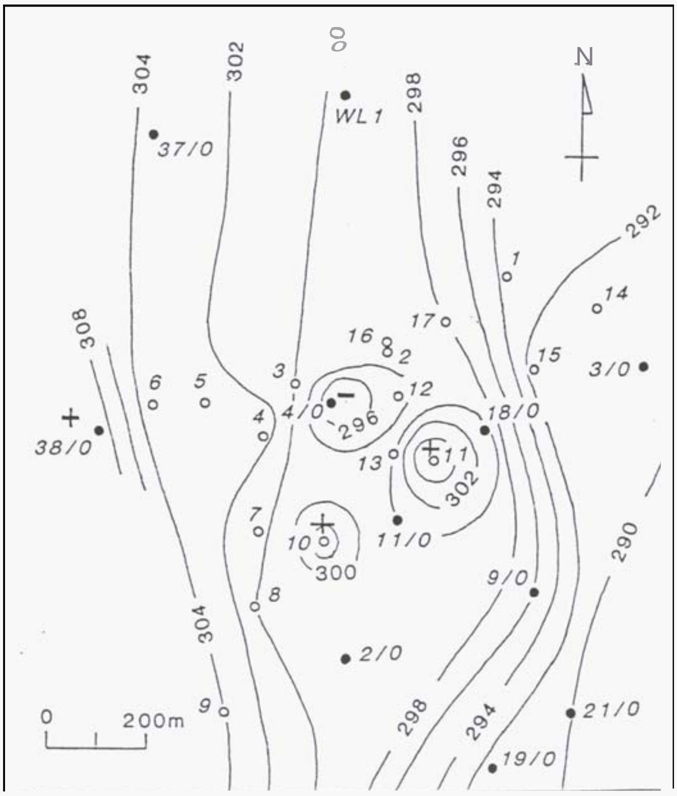

Fig. 5 Contours of groundwater level surface (m; R.L.) on 22/4/93, inferred from measurements in newly drilled "WLB" boreholes (O) and existing monitor holes (●).

are still elevated. The large lateral gradients and rapid changes in apparent water level in this area suggest that these wells are still controlled by perched aquifers which drain into the regional groundwater aquifer, and are recharged by condensed steam and rainfall. The cone of depression in the regional groundwater around BR4/0 reduces to about 1 to 2 m at about 100 m radius. We speculate that the depression is centred on a collapse pit, the site of an almost extinct fumarole, about 30 m east of BR4/0. The base of this pit is at approximately the same level (306 m R.L.) as the locally perched aquifer which controlled the water level in BR4/0 prior to exploitation.

Similar collapse pits or craters associated with active or extinct thermal features are found elsewhere in the north-western part of the Ohaaki field. It is likely that they are located above permeable conduits such as faults, or ancient hydrothermal eruption vent pipes (Hochstein & Henrys, 1989), where steam has invaded and heated the groundwater. Other groundwater monitor bores (including 2/0, 3/0, 9/0, 11/0) and Ohaaki Pool, which demonstrated a water level drop of several metres during the 1968-71 and 1988-93 field discharges, are all located near active or extinct thermal features. These drawdowns are also likely to be caused by local downflows of groundwater into the underlying geothermal aquifers, as deep reservoir pressures are drawn down.

6. CONCLUSIONS

- GPR, VES, and SP measurements have limited ability to determine the depth, and changes in depth, of shallow groundwater in the Ohaaki field; shallow monitor bores are probably the best tool.

- The original concept of one large cone of groundwater depression, centred on BR4, should be replaced by the concept of several small localised depressions (about 5 m), probably associated with near extinct thermal features.

7. ACKNOWLEDGEMENTS

We acknowledge funding from the Foundation for Science Research and Technology, and Waikato Regional Council for this work. Electricity Corporation of New Zealand provided material support.

8. REFERENCES

Alcaraz, A.P.; Barker, B.J.; Datuin, R.T.; Powell, T.S. (1989). The Tiwi field: a case study of geothermal development for the national interest. Proceedings 11th NZ Geothermal Workshop: 261-265.

Allis, R.G. (1980). Possible effects of reinjection at Wairakei Geothermal Field. Geothermal Resources Council Transactions, 4: 389-392.

__________ (1982). Geologic controls on shallow hydrologic changes at Wairakei field. Proceedings 4th NZ Geothermal Workshop: 139-144.

Bixley, P.F. (1990). Cold water invasion in producing liquid dominated geothermal reservoirs. Proceedings 15th Workshop on Geothermal Reservoir Engineering, Stanford University: 187-191.

Electricorp (1990). Water Right Application and Impact Assessment: Wairakei Geothermal Power Station. Electricity Corporation of New Zealand.

Grant, M.A. (1981). The effect of cold water entry into a liquid-dominated two-phase geothermal reservoir. Water Resources Research, 17: 1033-1043.

__________ (1982). Recharge to the Wairakei reservoir. Proceedings 4th NZ Geothermal Workshop: 33-37.

Grant, M.A.; Donaldson, I.G.; Bixley, P.F. (1982). "Geothermal Reservoir Engineering", Academic Press, London.

Hochstein, M.P.; Mayhew, I.D.; Villarosa, R.A. (1990). Self-potential surveys of the Mokai and Rotokawa high temperature field (New Zealand). Proceedings 12th NZ Geothermal Workshop, Auckland University.

Hochstein, M.P. and Henrys, S.A. (1989). Geophysical structure and densification of producing layers in the Broadlands-Ohaaki field (New Zealand). Proceedings 11th NZ Geothermal Workshop, Auckland University.

Hunt, T.M. (1987). Recharge at Broadlands Geothermal Field 1967-1983 determined from repeat gravity measurements. Research Report 216, Geophysics Division, DSIR, Wellington, New Zealand.

Hunt, T.M. and Bibby, H.M. (1992). Geothermal hydrology in "Waters of New Zealand" (P Moseley, ed.) NZ Hydrological Society.

Sherburn, S.; Clotworthy, A.; Hunt, T.M. (1993). No induced seismic activity at Ohaaki Geothermal Field. Proceedings 15th NZ Geothermal Workshop: this volume.

Villadolid, F.L. (1991). The applications of natural tracers in geothermal development: the Bulalo, Philippines experience. Proceedings 13th NZ Geothermal Workshop: 69-74.